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# Accreted seamounts in North Tianshan, NW China: Implications for the evolution of the Central Asian Orogenic Belt

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## ABSTRACT

The Carboniferous Bayingou ophiolitic *mélange* is exposed in the North Tianshan accretionary complex in the southwestern part of the Central Asian Orogenic Belt (CAOB). The *mélange* is mainly composed of serpentinised ultramafic rocks (including harzburgite, lherzolite, pyroxenite, dunite and peridotite), pillowed and massive basalts, layered gabbros, radiolarian cherts, pelagic limestones, breccias and tuffs, and displays block-in-matrix structures. The blocks of ultramafic rocks, gabbros, basalts, cherts, and limestones are set in a matrix of serpentinised ultramafic rocks, massive basalts and tuffs. The basaltic rocks in the *mélange* show significant geochemical heterogeneity, and two compositional groups, one ocean island basalt-like, and the other mid-ocean ridge-like, can be distinguished on the basis of their isotopic compositions and immobile trace element contents (such as light rare earth element enrichment in the former, but depletion in the latter). The more-enriched basaltic rocks are interpreted as remnants/fragments of seamounts, derived from a deep mantle reservoir with low degrees (2–3%) of garnet lherzolite mantle melting. The depleted basalts most likely formed by melting of a shallower spinel lherzolite mantle source with ~15% partial melting. It is probable that both groups owe their origin to melting of a mixture between plume and depleted MORB mantle. The results from this study, when integrated with previous work, indicate that the Junggar Ocean crust (comprising a significant number of seamounts) was likely to have been subducted southward beneath the Yili-Central Tianshan block in the Late Devonian-Early Carboniferous. The seamounts were scraped-off and accreted along with the oceanic crust in an accretionary wedge to form the Bayingou ophiolitic *mélange*. We present a model for the tectonomagmatic evolution of this portion of the CAO involving prolonged intra-oceanic subduction with seamount accretion

## 1. Introduction

Seamounts are one of the most ubiquitous landforms on Earth and at the present-day are distributed on all major oceanic plates (Wessel et al., 2010). Indeed, it has been reported that about 12,000 seamounts larger than 1.5 km in height can be found on the floor of the current oceans (Watts et al., 2010). Although seamounts can help us to better understand various deep Earth processes (e.g., Koppers and Watts, 2010), they are one of the least well-understood types of magmatic province and significant questions relating to their origin and structure remain (Foulger and Natland, 2003; Staudigel and Clague, 2010). Seamounts are generally interpreted to have formed by upwelling and decompression melting of mantle plumes (Morgan, 1971; Regelous et al., 2003) and thus can be recorders of both the speed and direction of plate motions (Tarduno et al., 2003). However, some seamounts form along lithospheric faults in response to plate flexure both during subduction (Hirano et al., 2006) and well away from plate boundaries (Machida et al., 2015).

Regardless of how they are formed seamounts migrate with the ocean plates and eventually reach active convergent margins (von Huene, 2008; Staudigel et al., 2010). However, because they represent areas of thicker-than-normal oceanic crust, they often accrete on the margin of subduction zones (Hoernle et al., 2002; Buchs et al., 2011). The attempted subduction of seamounts has also been linked to earthquake generation (Kodaira et al., 2000; Singh et al., 2011), submarine landslides (Ruh, 2016), as well as crustal erosion at subduction zones (Ranero and van Huene, 2000). The accretion of seamounts during subduction (e.g., Hoernle et al., 2002; Buchs et al., 2011; Prendergast and Offler, 2012), means that they can be preserved in the geological record, and thus can provide important insights into magmatic and mantle processes in long-destroyed oceans. They also can provide insights into the processes of continental growth at destructive plate margins (Buchs et al., 2016).

The Altaids or Central Asian Orogenic Belt (CAOB) represents one of

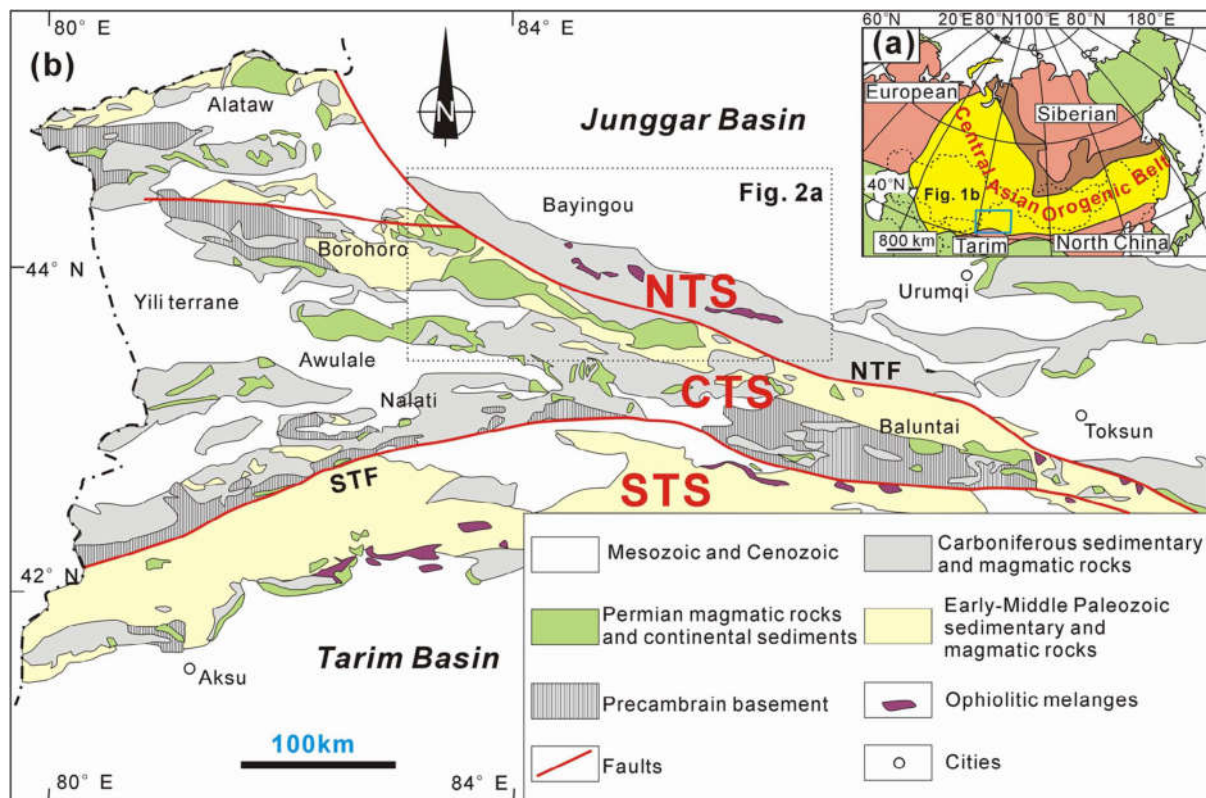


Fig. 1. (a) Overview map showing the Central Asian Orogenic Belt and adjacent cratons. (b) The tectonic geological map of the Chinese Tian Shan. NTS=North Tianshan accretionary complex; CTS=Yili-Central Tianshan block; STS= South Tianshan accretionary complex (modified after Liu et al., 2015).

the largest accretionary orogens in the world and comprises subduction- accretion complexes, arc rocks, ophiolites, seamounts and microcontinents accreted from the Neoproterozoic to the late Phanerozoic, during the closure of the paleo-Asian Ocean (Fig. 1a; Şengör et al., 1993; Windley et al., 2007; Wilhem et al., 2012; Xiao et al., 2015; Yang et al., 2015). The vast accretionary complexes of the CAOB contain a significant proportion of more-enriched basalts (Yang et al., 2015), similar in composition to modern-day ocean island basalts (OIB). On the basis of their geochemistry and nature of associated sediments many of these enriched basalts have been interpreted as the accreted remnants of seamounts (e.g., Safonova and Santosh, 2014; Yang et al., 2012a, 2012b, 2013, 2015). However, since accreted seamounts generally get caught up in ophiolitic mélangé belts they have undergone intense deformation, so can easily be misinterpreted as “classic” ophiolitic units (cf. Khan et al. (2007) and Kakar et al. (2014)). The exceptions are the large seamounts (like the Louisville and Emperor seamounts), which are more likely to be accreted rather than partially subducted, these are, however, rather rare in the geological record (Hauri, 1996). In this paper we present new tectonostratigraphic and geochemical data from accreted seamounts exposed in the North Tianshan accretionary complex (Xiao et al., 2013). Our observations provide convincing evidence for seamount development in the paleo- Asian Ocean, and represent a new opportunity for the study of remnant oceanic intraplate volcanoes in the CAOB.

## 2. Geological background and features of the Bayingou ophiolitic mélangé

As the southern branch of the CAOB, the Tianshan orogenic belt is located between the Junggar and Tarim Basins, and extends E-W about 3000 km from NW China to Kazakhstan

and Kyrgyzstan (Fig. 1b). The Chinese part of the Tianshan orogenic belt is generally divided into eastern and western segments roughly along longitude 88°E (Li et al., 2006). Tectonically, the Western Tianshan orogen can be divided into North Tianshan accretionary complex (NTAC), Yili-Central Tianshan block (CTB) and South Tianshan accretionary complex (STAC) that are bounded by the North Tianshan fault (NTF) and the South Tianshan fault (STF), respectively (Fig. 1b).

The NTAC is located between the Junggar terrane and Yili-Central Tianshan block in the SW of the CAO (Fig. 2a; Zhu et al., 2005; Gao et al., 2009b; Wang et al., 2009; Xiao and Santosh, 2014), and extends over 300 km along the northern edge of the CTB. The NTAC is bounded in the south by the NW–striking North Tianshan fault and in the north by a north-directed thrust that has thrust the accretionary complex over the Permian and Cretaceous sequences. The accreted sequence predominantly comprises two lithologies: (1) Devonian–Carboniferous volcanosedimentary rocks widely regarded to be an accretionary complex (Wang et al., 2006), and (2) ophiolitic remnants (Gao et al., 1998). The siliceous rocks in the accretionary complex contain Famennian (372–359 Ma) conodonts and Lower Carboniferous radiolarians (Xiao et al., 1992), while gabbro from the Bayingou ophiolitic mélange has



yielded zircon U-Pb ages of  $344 \pm 3$  Ma (Xu et al., 2006a).

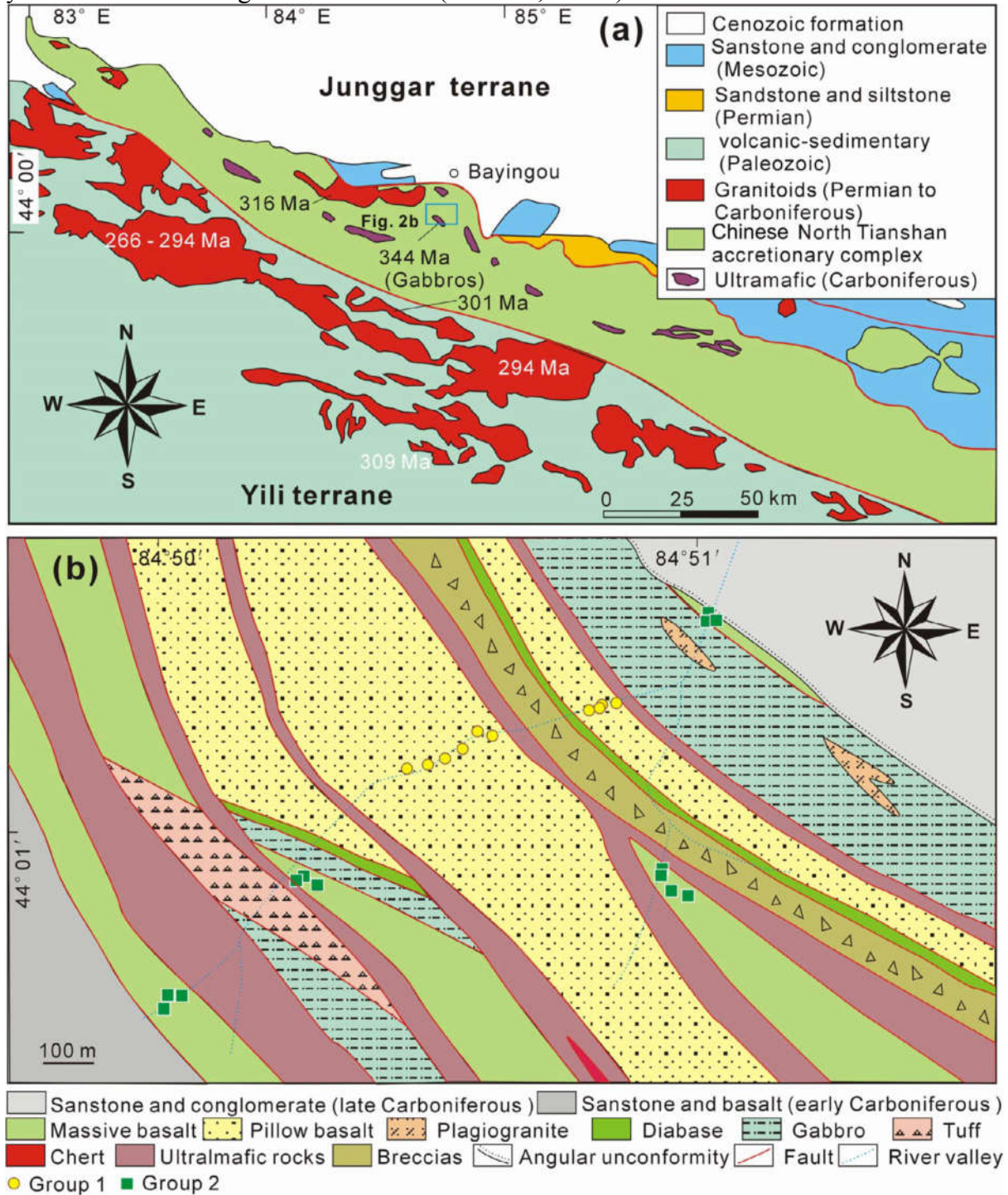


Fig. 2. (a) Tectonic map showing the North Tianshan accretionary complex and adjacent region (modified after Han et al., 2010). (b) Geological map of the Bayingou ophiolitic mélangé (modified after Xu et al., 2006b) and sample locations.

The undeformed Sikeshe granodiorite, K-feldspar granite and diorite pluton that intruded into flysch has been dated by SHRIMP zircon U-Pb at  $316 \pm 3$  Ma (Han et al., 2010).

The CTB is suggested to represent a composite arc terrane that developed on a Precambrian basement (Gao et al., 2009a), which consists of the Precambrian amphibolite- and granulite-facies metamorphic rocks (Liu et al., 2004; Shu et al., 2004). Ordovician–Silurian



metavolcanic-sedimentary rocks are fault contact with the Precambrian basement, and are overlain unconformably by non-metamorphosed the Lower Carboniferous sedimentary rocks (Gao et al., 1998; Charvet et al., 2011). The volcanic rocks interbedded within the Paleozoic sediments have subduction-related geochemical features (Che et al., 1994; Zhu et al., 2005, 2009; Li et al., 2010, 2017). The granitoids, which are widely distributed and intruded into the Proterozoic to Paleozoic volcanic-sedimentary rocks, are mainly composed of adamellite, moyite, diorite and quartz diorite with different ages (Tang et al., 2010; Long et al., 2011; Dong et al., 2011; Ma et al., 2014; Wang et al., 2015; Zhong et al., 2015). Moreover, several ophiolitic mélanges have

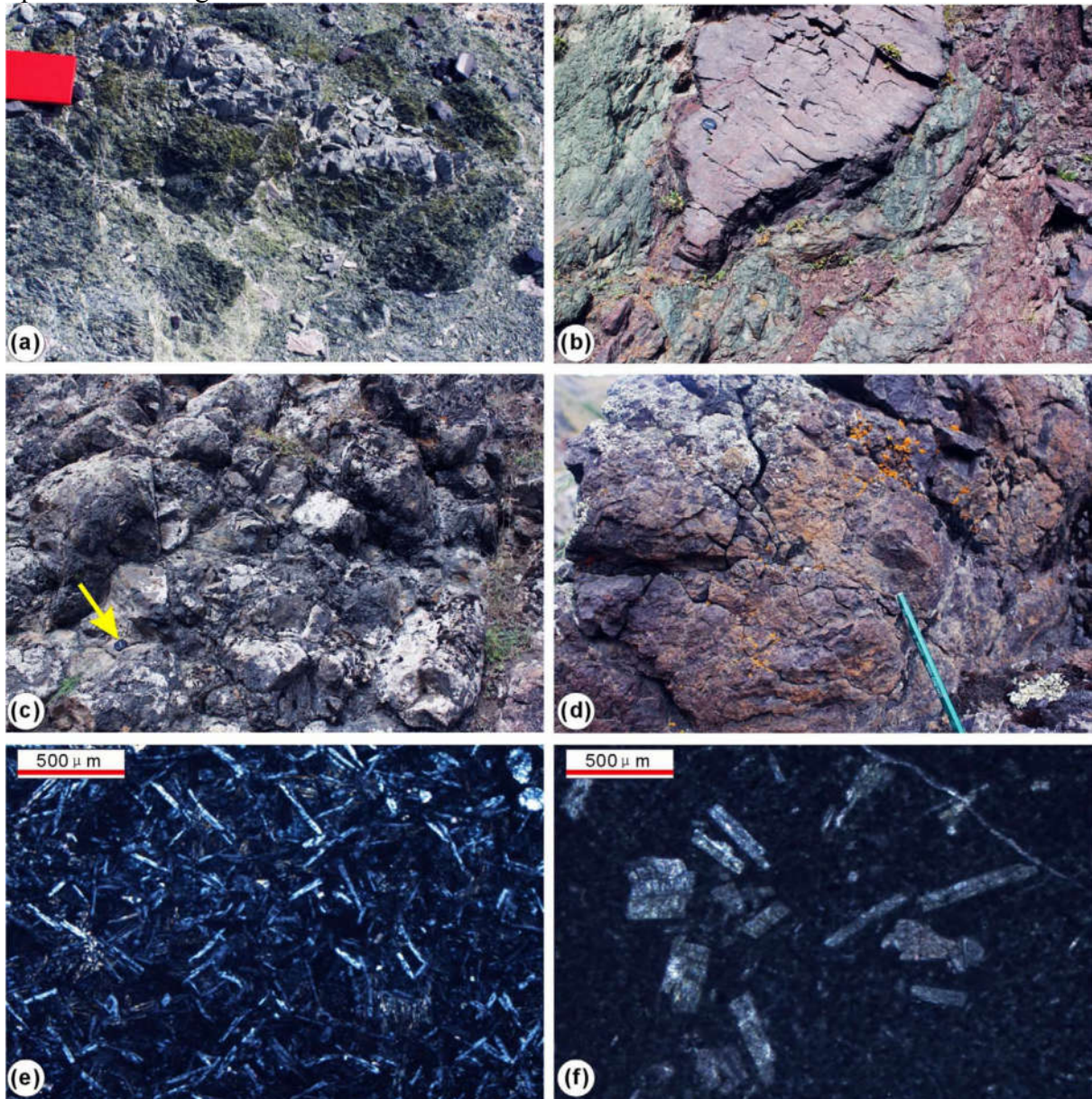


Fig. 3. Field photographs of typical outcrops and photomicrographs from the Bayingou ophiolitic mélangé in the North Tianshan accretionary complex. (a) Metagabbros in ultramafic rocks with cleavages. (b) Chert lens occurring in tuffs. (c) Pillow basalts from Group 1 with interstratified tuffs. (d) Massive basalts in the field (Group 2). (e) Equigranular textures in pillow basalts (cross-polarised light, group 1). (f) Typical porphyritic texture in massive basalts (cross-polarised light, group 2).

been recognized in the CTB, such as the Bingdaban, Mishigou and Gangou ophiolitic mélanges (Charvet et al., 2007; Dong et al., 2006, 2007).

The STAC marks the suture between the Yili-Central Tianshan block and the Tarim Basin (Xiao et al., 2013; Jiang et al., 2014), and consists predominantly of the Lower Cambrian black shales and phosphoric silicates, Cambrian–Carboniferous marine/non-marine carbonates, clastic rocks, cherts and interlayered volcanics (Allen et al., 1992; Carroll et al., 1995). Several ophiolitic mélanges with ages ranging from 450 Ma to 332 Ma developed in the STAC (Gao et al., 1998; Wang et al., 2007; Jiang et al., 2014). Subduction- and/or collision-related high-pressure/low-temperature metamorphism is also preserved (Gao et al., 1998; Zheng et al., 2006; Lü et al., 2008). Moreover, large volume Late Carboniferous to Permian syn- and post-collisional granitic plutons are found intruding the STAC (Konopelko et al., 2009; Long et al., 2011; Han et al., 2011; Gou et al., 2012).

The Bayingou ophiolitic mélange is exposed discontinuously over an area ~250 km in length by 5–15 km in width (Fig. 2b), and consists of strongly serpentinised ultramafic rocks (including harzburgite, lherzolite, pyroxenite, dunite and peridotite), as well as pillowed and massive basalts, layered gabbros, radiolarian cherts, pelagic limestones, breccias and tuffs. These rocks show a series of tight isoclinal folds and display block-in-matrix structures (Figs. 3a and b). The blocks range from tens of centimeters to several hundreds of meters and are set in a matrix of serpentinised ultramafic rocks, basalts and tuffs. In most cases the matrix is strongly sheared so that a scaly fabric is developed in weakly metamorphosed pelite. Harzburgite, one of the main ultramafic rocks, contains more than 70% olivine, 15% clinopyroxene, 5% orthopyroxene, and about 5% spinel. Dunite, another main ultramafic rock, contains more than 95% olivine, 5% of orthopyroxene. However, both ultramafic rocks are strongly serpentinised with a net-like texture of chrysotile and antigorite with only a few relicts of fresh olivine and enstatite being preserved.

Gabbro occurs mainly as blocks but small sheet-like layered intrusions are preserved in some places, and mainly consist of clinopyroxene and plagioclase, as well as ilmenite, magnetite, sphene

**Table 1**  
Major element compositions of basalts from the Bayingou ophiolitic mélange.

Sample	Type	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	Fe <sub>2</sub> O <sub>3</sub>	MnO	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O	P <sub>2</sub> O <sub>5</sub>	LOI	Total
BY-1	Group 1	50.58	2.23	14.46	10.30	0.13	6.76	8.97	5.15	0.13	0.34	0.90	99.95
BY-2		50.17	2.77	14.82	10.87	0.12	5.59	8.39	6.00	0.13	0.40	0.60	99.86
BY-3		51.58	2.53	13.56	9.90	0.17	6.96	7.62	5.74	0.15	0.52	1.10	99.83
BY-4		49.39	2.89	13.52	11.12	0.21	6.87	8.15	5.57	0.15	0.43	1.24	99.54
BY-5		49.56	2.89	13.78	11.21	0.15	6.82	7.95	5.65	0.16	0.54	1.08	99.79
BY-6		51.12	2.76	14.49	9.87	0.22	6.17	7.86	5.57	0.15	0.53	1.08	99.82
BY-7		51.32	2.44	14.21	10.32	0.19	6.23	8.34	5.21	0.14	0.41	1.12	99.93
BY-8		50.72	2.68	13.89	11.13	0.17	6.15	8.14	5.28	0.16	0.48	1.21	100.01
BY-9		49.38	2.53	14.77	10.30	0.16	6.76	8.97	5.15	0.15	0.47	1.25	99.89
BY-10		50.22	2.81	14.22	10.62	0.19	5.78	8.18	5.89	0.16	0.47	1.18	99.72
BY-11	Group 2	49.96	1.33	13.97	15.53	0.24	4.79	6.70	5.14	0.70	0.20	1.37	99.93
BY-12		49.00	0.84	14.61	11.21	0.16	8.85	10.44	2.65	0.54	0.07	1.44	99.81
BY-13		50.59	0.88	14.01	11.35	0.18	9.01	9.99	3.65	0.10	0.07	0.50	100.33
BY-14		49.55	0.84	14.13	11.22	0.16	8.79	10.87	3.08	0.29	0.07	0.94	99.94
BY-15		49.97	1.68	13.10	16.88	0.18	4.29	5.96	5.99	0.45	0.28	0.95	99.73
BY-16		50.97	1.23	15.60	13.41	0.17	4.24	6.17	6.79	0.12	0.13	0.87	99.70
BY-17		50.46	1.25	15.37	13.55	0.17	4.99	5.92	6.46	0.48	0.18	1.12	99.95
BY-18		49.07	1.19	16.78	13.16	0.28	5.25	6.59	5.23	2.03	0.31	0.22	100.11
BY-19		48.19	1.13	16.00	11.52	0.22	6.37	10.39	3.34	1.89	0.11	0.83	99.99
BY-20		51.54	1.41	14.22	9.78	0.24	5.48	10.08	5.13	1.00	0.10	0.87	99.85
BY-21		50.62	1.98	14.07	14.52	0.21	6.00	5.07	5.89	0.41	0.28	0.65	99.70
BY-22		45.14	1.13	13.32	14.46	0.22	6.22	8.99	9.73	0.44	0.22	0.23	100.10
BY-23		50.50	1.53	13.90	13.84	0.21	5.61	5.96	6.76	1.04	0.15	0.61	100.11

and zircon. Plagioclase is usually enclosed by clinopyroxene which has a poikilitic texture. Gabbros have undergone low temperature metamorphism to hornblende, actinolite, epidote and chlorite. The basalts can be divided into two types: pillowed (Fig. 3c and e) and massive (Fig. 3d and f). The pillowed basalt occurs extensively as blocks in the Bayingou area, and



contains amygdalae that are filled with chalcedony or calcite. The pillowed basalts also display a porphyritic texture with clinopyroxene and plagioclase phenocrysts in groundmass of fine-grained plagioclase, clinopyroxene and opaque minerals. The groundmass plagioclase has a quench texture which is a unique feature of marine volcanic rocks (Peng et al., 2005). The massive basalt occurs as not only matrix but also blocks in the *mélange*, and shows an interstitial texture as well as a porphyritic texture. These basalts are composed of plagioclase, clinopyroxene, magnetite, ilmenite, and minor zircon, and also show low-grade metamorphism with albite, epidote, chlorite, and calcite. In previous regional studies, the mafic rocks of the Bayingou ophiolitic *mélange* were suggested to belong to three different geochemical groups: N-MORB, OIB and island arc tholeiite (IAT) (Wu and Liu, 1989; Xiao et al., 1992; Xu et al., 2006b; Chen et al., 2012). Chert is generally juxtaposed against massive and pillowed basalt, and has undergone multiple folding and thrusting. Bedded pelagic limestone is found along with massive and pillowed basalt, chert, and tuff, and occurs mostly as blocks in a matrix of turbidite.

### 3. Analytical methods

In order to determine the composition of the seamounts, 23 whole-rock samples of basalts were selected for X-ray fluorescence (XRF) and inductively coupled plasma–mass spectrometry (ICP-MS) analyses at the Chang'an University (China). Analytical procedures are similar to those described by Li et al. (2006). Based on the analysis of international reference materials and duplicate samples, the analytical errors for the major elements were about 1–2 wt.% for SiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, TiO<sub>2</sub>, Fe<sub>2</sub>O<sub>3</sub> and MgO and less than 3 wt.% for the other oxides. About 50 mg of each powdered sample was dissolved in a high-pressure Teflon bomb for 24 h using a HF+ HNO<sub>3</sub> mixture. International reference material (BE-N) was used to monitor data quality during the course of this study. The precision and accuracy of the trace element analyses are estimated to be better than 5 wt.% (relative), except for Nb and Ta (better than 10 wt.%). For detailed information see appendix Table 1.

The isotope ratios of Nd, Sr and Pb along with associated isotope–dilution concentrations were measured at the Test and Research Center, Beijing Nuclear Industry Geological Research Institute, China. <sup>87</sup>Sr/<sup>86</sup>Sr and <sup>143</sup>Nd/<sup>144</sup>Nd ratios were measured on a MAT-261 mass spectrometer, and isotopic ratios were normalised to <sup>146</sup>Nd/<sup>144</sup>Nd = 0.7219 and <sup>86</sup>Sr/<sup>88</sup>Sr = 0.1194. Repeat analyses yielded averages of 0.710236 ± 0.000007 (2σ, n = 6) for Sr standard NIST SRM987, and 0.511864 ± 0.000003 (2σ, n = 6) for the La Jolla Nd standard. Total elemental blanks were <200 pg for Sr and <100 pg for Nd. For whole rock Pb isotopic compositions, about 100 mg powder was accurately weighed into a Teflon beaker, spiked and digested in concentrated HF at 180 °C for 7 h. Lead was separated and purified by using AG1×8, 20–400 cation-exchange resin with dilute HBr as an eluant. Total procedural blanks were <50 pg Pb. Lead isotopic ratios were measured using a VG-354 mass spectrometer. Repeat analyses of SRM 981 yielded average values of <sup>206</sup>Pb/<sup>204</sup>Pb = 16.9325 ± 3 (2σ), <sup>207</sup>Pb/<sup>204</sup>Pb = 15.4853 ± 3 (2σ) and <sup>208</sup>Pb/<sup>204</sup>Pb = 36.6780 ± 9 (2σ). External precision for the Pb isotope systems are less than 0.005, 0.005 and 0.0015, respectively.

### 4. Results

As detailed in Section 2, most of the samples are quite altered (principally to chlorite and epidote) and so have undergone significant post-magmatic fluid-rock interaction. However, they have relatively low loss on ignition values (LOI < 1.37 wt.%; Table 1). Despite these low LOI values hydrothermal alteration is likely to have mobilised some of the major elements and the large ion lithophile (LIL) elements (Ba, Rb, Sr, and K), as has been



documented in numerous studies (e.g., Pearce, 1996; Polat and Hofmann, 2003). However, the high field- strength (HFS) elements such as Nb, Ta, Zr, Hf, Y and Th are generally regarded to be relatively immobile during low-temperature alteration (e.g., Pearce and Cann, 1973; Polat et al., 2002). Although Sr isotopes can be significantly affected by interaction with high  $^{87}\text{Sr}/^{86}\text{Sr}$  seawater (e.g., Kawahata et al., 2001; Godard et al., 2006), Nd and Pb radiogenic isotope ratios are relatively insensitive to alteration and are much more reliable indicators of magmatic sources (Rodrigo-Gámiz et al., 2015; Simon et al., 2017). We therefore use immobile trace elements along with Nd and Pb radiogenic isotope ratios as well as field observations and regional constraints in our assessment of the petrogenesis and tectonic setting of these rocks.

Two distinct compositional groups are evident within the geochemically heterogeneous Bayingou ophiolitic mélange (Fig. 4). Group 1 have MgO ranging from 5.6 to 7.0 wt.% and are characterised by high

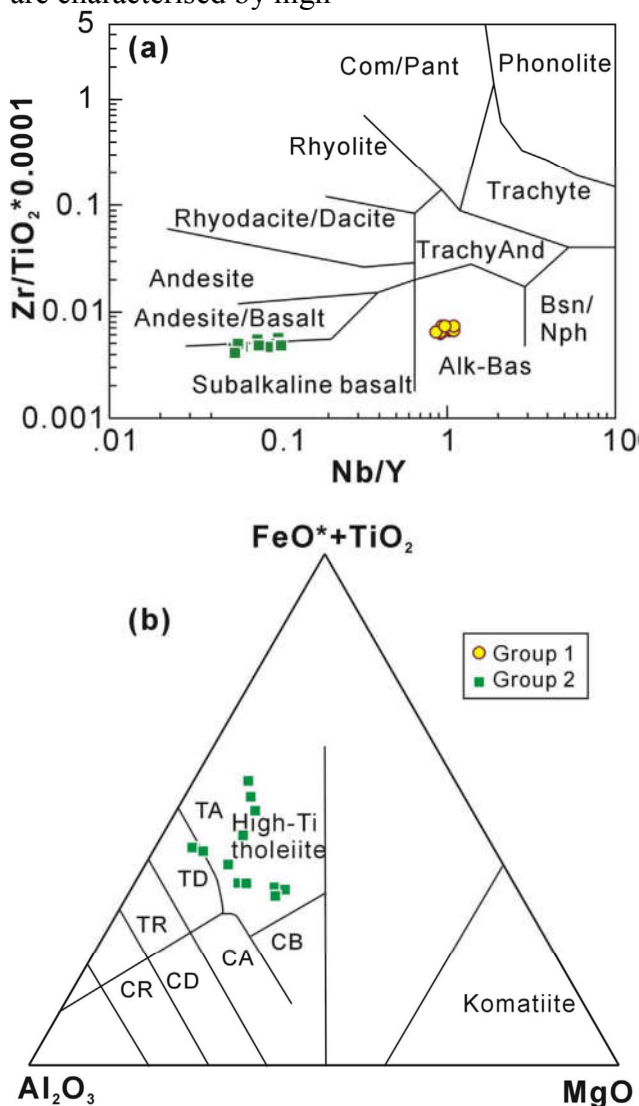


Fig. 4. Geochemical classification diagrams for the basaltic units in the Bayingou ophiolitic mélange in the North Tianshan accretionary complex. (a)  $\text{Zr}/\text{TiO}_2$  versus  $\text{Nb}/\text{Y}$  (after Winchester and Floyd, 1977). (b)  $\text{Al}_2\text{O}_3$ – $\text{FeO} + \text{TiO}_2$ – $\text{MgO}$  diagram (Jensen, 1976); tholeiitic series: TA - andesite, TD - dacite, TR - rhyolite; calc-alkaline series: CB - basalt, CA - andesite, CD - dacite, CR - rhyolite; kom - komatiite.

concentrations of TiO<sub>2</sub> (2.2–2.9 wt.%, Table 1). Magnesium number (Mg# = molar Mg<sup>2+</sup>/[Mg<sup>2+</sup>+Fe<sup>2+</sup>], assuming 85% of total Fe is Fe<sup>2+</sup>) varies from 51 to 59. Group 1 samples have a restricted compositional range, and do not show a clear linear relationship with increasing MgO (Fig. 5). In contrast, P<sub>2</sub>O<sub>5</sub>, Cr and Ni in Group 1 samples range to higher values than in Group 2 (Table 2). However, Al<sub>2</sub>O<sub>3</sub> range to lower than in Group 2 (Fig. 5).

These rocks classify as alkaline basalts on an immobile element (Zr-Ti-Nb-Y) classification plot (Fig. 4a). This alkaline affinity is also evidenced by light rare earth element (LREE) enrichment and heavy rare earth element (HREE) depletion (Fig. 6a, (La/Yb)<sub>N</sub>=5.9–7.5, N - Chondrite normalised). Group 1 samples also show slight negative Zr-Hf anomalies and have a similar trace element signature to ocean island basalt (OIB) (Fig. 6b, Fig. 7). The rocks of this group have a very restricted range of Nd-Pb isotopic compositions ( $\epsilon$ Nd(t)=+5.8–+7.0, <sup>206</sup>Pb/<sup>204</sup>Pb=19.39–19.53; <sup>207</sup>Pb/<sup>204</sup>Pb=15.71–15.73; <sup>208</sup>Pb/<sup>204</sup>Pb=40.03–40.21; Tables 3 and 4). The basalts have lower  $\epsilon$ Nd(t) and <sup>208</sup>Pb/<sup>204</sup>Pb for a given <sup>206</sup>Pb/<sup>204</sup>Pb, than the Society and Cook-Austral islands and the Louisville Seamount Chain but overlap with the field for the Hawaiian basalts. The samples plot close to the DMM end member and seem to fall on a trend between depleted mantle and EM1 (Fig. 8). Although Sr isotopic compositions span a relatively small range (<sup>87</sup>Sr/<sup>86</sup>Sr=0.70493–0.70511), they do not look like magmatic values and are probably influenced by seawater alteration as mentioned above.

Group 2 basaltic rocks (4.3–9.0 wt.% MgO) have a much more incompatible element-depleted (MORB-like) signature (Fig. 6a) with LREE depletion (La/Sm)<sub>N</sub>= 0.5–0.9; N - Chondrite normalised) and essentially flat HREE patterns (Dy/Yb)<sub>N</sub>= 0.9–1.2; N - Chondrite normalised). They are also characterised by lower concentrations of TiO<sub>2</sub> (0.9–2.0 wt.%) and a greater variation in Mg# (34–62) than Group 1, again with no significant negative Nb, Ta or Ti anomalies (Fig. 6b). This group are much more tholeiitic in nature (Fig. 4b) and straddle the boundary between basaltic andesite and sub-alkaline basalt on an immobile element (Zr-Ti-Nb-Y) classification plot (Fig. 4a). This more tholeiitic nature is also consistent with the depleted MORB-like trace element signature (Fig. 7). Despite their trace element-depleted signature of Group 2 has slightly more enriched  $\epsilon$ Nd(t) (3.2–4.4; Table 3) and lower initial <sup>206</sup>Pb/<sup>204</sup>Pb (18.28–18.48), <sup>207</sup>Pb/<sup>204</sup>Pb (15.51–15.57), and <sup>208</sup>Pb/<sup>204</sup>Pb (38.43–38.60) compared to those of Group 1 (Fig. 8, Table 4). The spread in Sr isotope values along with a relatively small range in Nd and Pb implies that Sr isotopes have been affected by seawater alteration. These samples fall along the same trend between DMM and EM1 as Group 1 samples but, paradoxically, the more incompatible trace-element depleted Group 2 plot closer to the compositional field of EM1 than the more trace element-enriched Group 1 samples (Fig. 8).

## 5. Discussion

### 5.1. Source nature

As shown in Section 4, their geochemistry indicates that Group 1 have OIB-like trace element normalised patterns (Fig. 6) with (La/Yb)<sub>N</sub> ratios >6 (Table 2). Immobile element ratios and discrimination diagrams (Figs. 7 and 9) further suggest that they have formed in a within plate setting. The association of the basalts with marine sediments indicates they did not erupt in a continental setting, but rather formed in an ocean basin and so are probably the result of mantle plume-related magmatism. La/Sm (3.3–3.9) and Th/Yb (1.1–1.4) values for Group 1 are extremely high, similar to the Hawaii and Louisville basalts (Fig. 7a), suggesting

that they were generated through

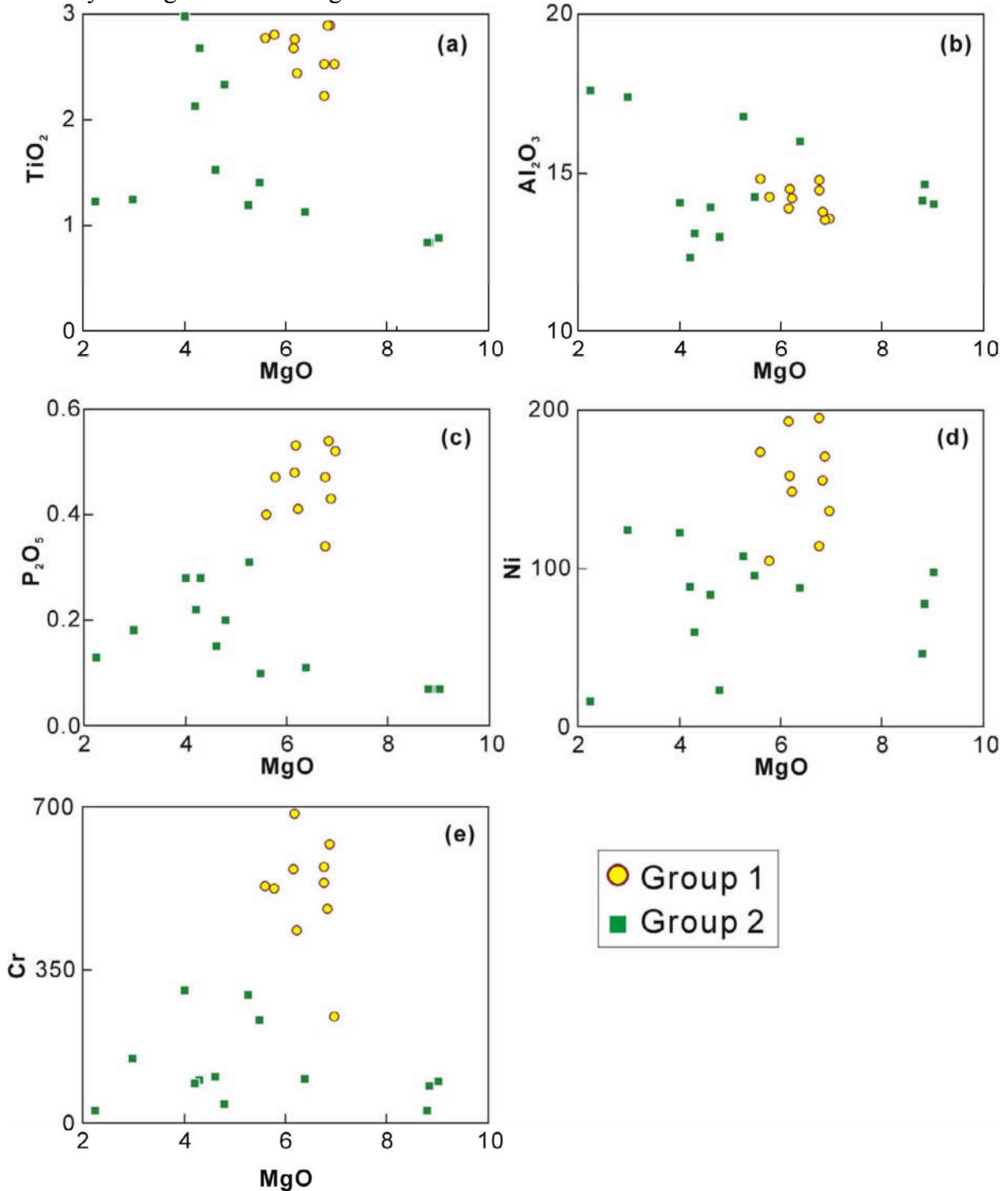


Fig. 5. Representative variation diagrams of major and trace element compositions versus MgO for the basalts in the Bayingou ophiolitic mélangé.

small degrees of melting of a mantle source. Samples in Group 1 are consistently slightly more enriched in LREE relative to Group 2. The geochemistry indicates that Group 2 have MORB-like trace element normalised patterns (Fig. 6) with  $(\text{La}/\text{Yb})_N$  ratios  $< 1$  (Table 2). However, immobile element ratios and discrimination diagrams (Figs. 7 and 9) suggest that they have formed in mid-oceanic ridge setting but interacted with a more-enriched plume source. Therefore, it is likely that both groups owe their origin to melting of a mixture between plume and depleted MORB mantle.



Pooled fractional melt calculations have been used to model the composition (depleted vs. primitive mantle) and mineralogy (lherzolite containing either spinel or garnet or a 50:50 mix representing melting across the transition zone) of the source region (Aldanmaz et al., 2006; Kerr et al., 2004, 2009). Partition coefficients are averages from Salters et al. (2002) and McDade et al. (2003). Mantle mineral proportions and source compositions have been taken from McKenzie and O'Nions (1991).

The melt modelling suggests that the higher (La/Nd)<sub>cn</sub> ratios of Group 1 are consistent with a slightly more enriched source region (Fig. 10a). Group 1 samples all plot between the 50:50 garnet–spinel lherzolite and spinel lherzolite melt model curves, indicating that the source contains some garnet. The modelling further suggests that Group 1 lavas formed by relatively small (2–5%) degrees of mantle melting. In contrast, although the basalts in Group 2 have a more enriched isotopic signature, their depleted LREE and flat-HREE signatures are better modelled using a depleted shallower spinel lherzolite mantle source composition, that has undergone 10–20% melting.

It is therefore likely that Group 2 basalts, with incompatible element ratios that are mostly similar to MORB (Fig. 9) yet have radiogenic isotope compositions that are more enriched than Group 1, are derived from a mantle source that has undergone a previous melting episode. Since this depletion is not recorded in the isotopic signatures it is clear that this previous melting event occurred shortly before the formation of the Group 2 basalts and does

not indicate a long-term source

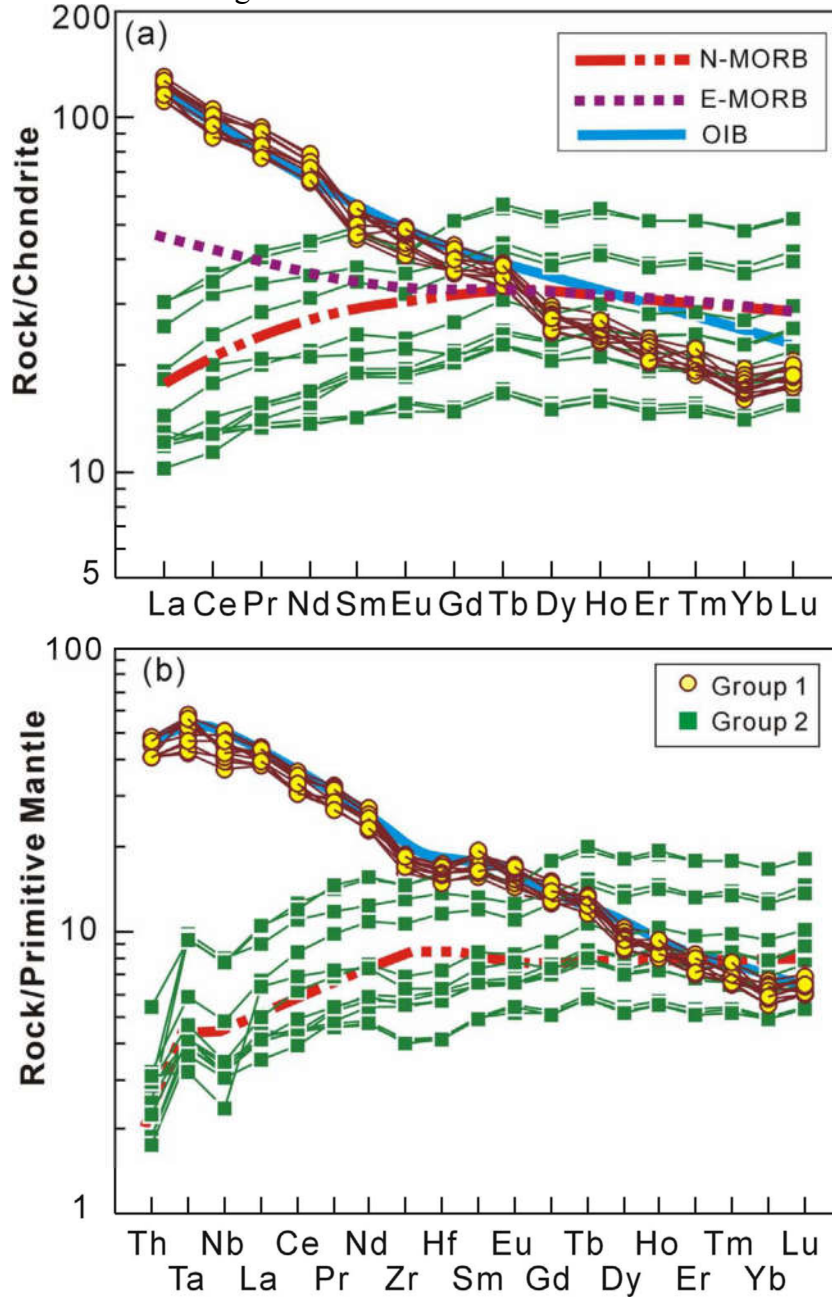


Fig. 6. (a) Chondrite-normalised rare earth element and (b) primitive mantle-normalised trace-element patterns for the basalts in the Bayingou ophiolitic mélange. The normalising values, E-MORB, OIB and N-MORB compositions are from Sun and McDonough (1989).

depletion. On all the isotope plots involving Pb and Nd, both groups of basalts lie on clear mixing trends between depleted MORB Mantle (DMM) and EM1, with Group 2 consistently plotting closer to EM1 than Group 1 (Fig. 8). The involvement of an enriched mantle source in the formation of Group 2 basalts (and to a lesser extent Group 1 basalts) along with the associated occurrence of pelagic sedimentary rocks support formation in an intra-oceanic setting. The fact that, both groups lie along a DMM-EM1 mixing line (Fig. 8) suggests that they may be genetically related.

## 5.2. Petrogenesis of basalts

It is generally assumed that the OIB are derived from mantle plumes and that these magmas commonly have more enriched compositions than MORB (Hofmann, 1997). Recent discoveries indicate that enriched basalts are not only limited to mantle plumes but can be found in oceanic spreading ridges (Niu et al., 2002; Hemond et al., 2006; Nauret et al., 2006), and in some arc-related settings (Kita et al., 2001; Scott et al., 2002) such as slab roll-back (Ferrari et al., 2001), arc rifting (Mueller et al., 2002) and deep water cycle model (Ivanov and Litasov, 2014). Moreover, Hirano et al. (2006) proposed that the small alkaline volcanoes erupt along lithospheric fractures in response to plate flexure

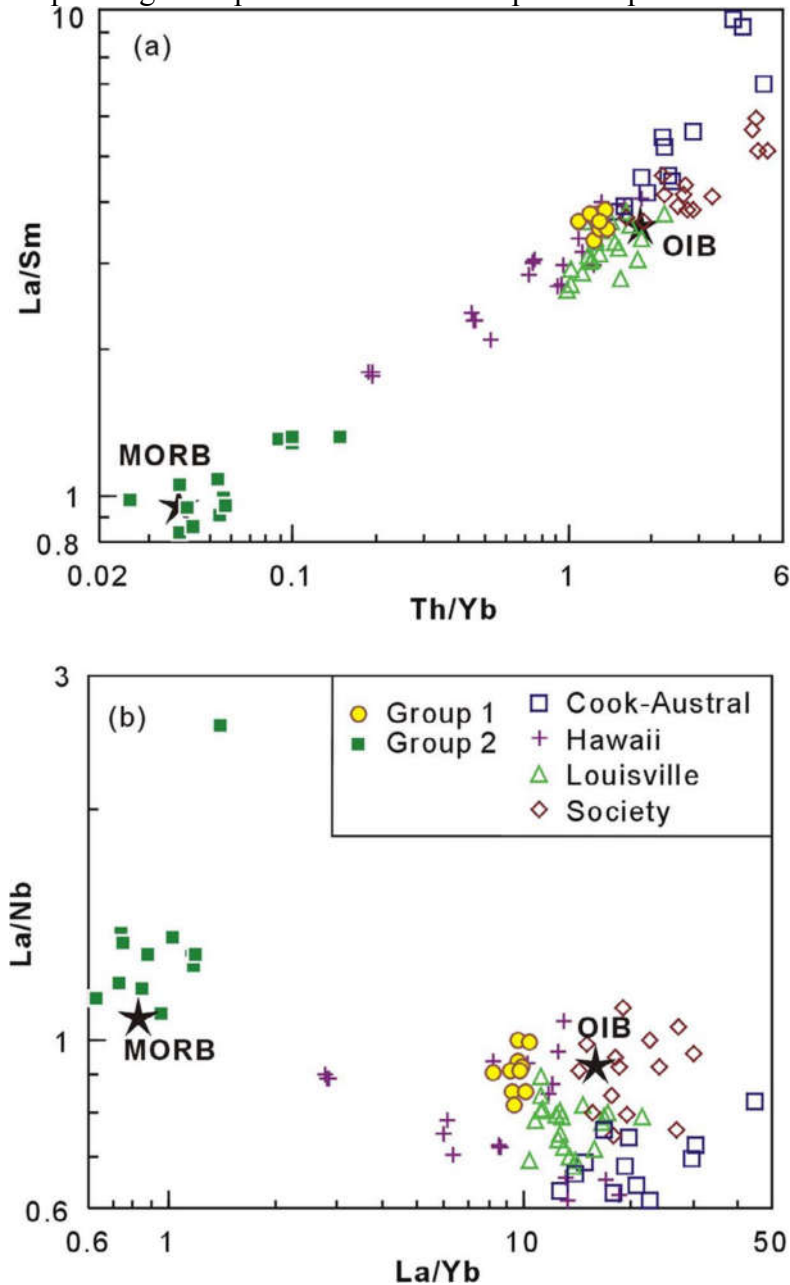


Fig. 7. Plots of (a) Th/Yb versus La/Sm, and (b) La/Yb versus La/Nb for the basalts in the Bayingou ophiolitic mélange. Data sources: Cook–Austral from Lassiter et al. (2003), Hawaii from Huang et al. (2005), Society from White and Duncan (1996), Louisville from Vanderkluyssen et al. (2014), OIB and MORB compositions are from Sun and McDonough (1989).

during subduction. In our previous studies we discussed the occurrences of alkaline basalts hosted by ophiolitic mélanges in the West Junggar and suggested that the plume-related magmatism developed during the evolution of the Junggar ocean as well as paleo-Asian



Ocean (Yang et al., 2012b, 2013, 2015). Safonova and Santosh (2014) also considered that the OIB-bearing units of the CAOB and the Western Pacific formed in relation to two superplumes: the Asian with age of the Late Neoproterozoic and the Pacific with age of the Cretaceous. Moreover, Fitton et al. (1997) have successfully developed the robust  $\Delta\text{Nb}$  approach ( $\Delta\text{Nb} = 1.74 + \log(\text{Nb}/\text{Y}) - 1.92 \times \log(\text{Zr}/\text{Y})$ ) to identify the mantle source of mafic rocks. Mafic magmas derived from plume exhibit  $\Delta\text{Nb} > 0$ , whereas magmas derived from depleted mantle and the crust show  $\Delta\text{Nb} < 0$  (Fitton et al., 1997; Baksi, 2001). The Group 1 basalts have  $\Delta\text{Nb}$  values ranging from 0.18 to 0.28, showing a geochemical signature (Table 2) consistent with a mantle plume origin. However, Group 2 basalts have  $\Delta\text{Nb}$  values ranging from 0.25 to 0.11, indicating mixing between plume and MORB mantle sources. This is compatible with both groups plotting on the DMM-EM1 mixing line (Fig. 8).

**Table 4**  
Pb isotope compositions of basalts from the Bayingou ophiolitic mélange in the North Tianshan.

Sample	Type	U	Th	Pb	$^{238}\text{U}/^{204}\text{Pb}$	$^{235}\text{U}/^{204}\text{Pb}$	$^{232}\text{Th}/^{204}\text{Pb}$	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$	$(^{206}\text{Pb}/^{204}\text{Pb})_i$	$(^{207}\text{Pb}/^{204}\text{Pb})_i$	$(^{208}\text{Pb}/^{204}\text{Pb})_i$
BY-1	Group 1	0.75	3.46	1.97	27.00	0.196	125.45	19.53	15.73	40.21	18.05	15.65	38.06
BY-2		0.91	4.1	2.1	30.73	0.223	139.46	19.39	15.71	40.04	17.71	15.62	37.64
BY-3		0.77	3.48	2.01	27.16	0.197	123.67	19.49	15.73	40.14	18.00	15.65	38.01
BY-4		0.73	3.51	1.92	26.96	0.196	130.58	19.39	15.72	40.16	17.91	15.64	37.91
BY-5		0.82	3.49	2.05	28.36	0.206	121.60	19.51	15.72	40.13	17.96	15.64	38.04
BY-6		0.83	4.05	1.96	30.03	0.218	147.59	19.49	15.71	40.03	17.84	15.63	37.50
BY-7		0.91	4.1	1.98	32.59	0.236	147.91	19.44	15.73	40.20	17.65	15.63	37.66
BY-8		0.78	3.98	2.03	27.25	0.198	140.04	19.53	15.72	40.20	18.03	15.64	37.79
BY-9		0.82	3.47	2.11	27.56	0.200	117.47	19.45	15.72	40.15	17.94	15.64	38.13
BY-10		0.88	3.87	1.96	31.84	0.231	141.03	19.40	15.72	40.16	17.66	15.62	37.74
BY-11	Group 2	0.07	0.25	1.45	3.23	0.023	12.32	18.43	15.52	38.56	18.25	15.51	38.35
BY-12		0.052	0.22	0.92	4.03	0.029	17.06	18.48	15.52	38.60	18.26	15.51	38.31
BY-13		0.069	0.24	0.31	15.72	0.114	55.70	18.48	15.51	38.44	17.62	15.46	37.48
BY-14		0.058	0.24	0.29	14.07	0.102	57.69	18.28	15.52	38.43	17.51	15.48	37.43
BY-15		0.13	0.21	0.69	12.89	0.093	21.74	18.48	15.57	38.53	17.77	15.53	38.16
BY-16		0.16	0.16	1.01	11.24	0.082	11.57	18.45	15.51	38.44	17.83	15.47	38.24
BY-17		0.17	0.16	0.87	13.74	0.100	13.42	18.36	15.53	38.46	17.60	15.49	38.23
BY-18		0.12	0.15	0.65	12.86	0.093	16.51	18.36	15.56	38.57	17.66	15.52	38.28

The unit for the contents of trace elements (U, Th, and Pb) is  $10^{-6}$ . The isotopic ratios  $^{238}\text{U}/^{204}\text{Pb}$ ,  $^{235}\text{U}/^{204}\text{Pb}$ , and  $^{232}\text{Th}/^{204}\text{Pb}$  are calculated from the measured whole rock U, Th, and Pb contents and the present Pb isotopic compositions.

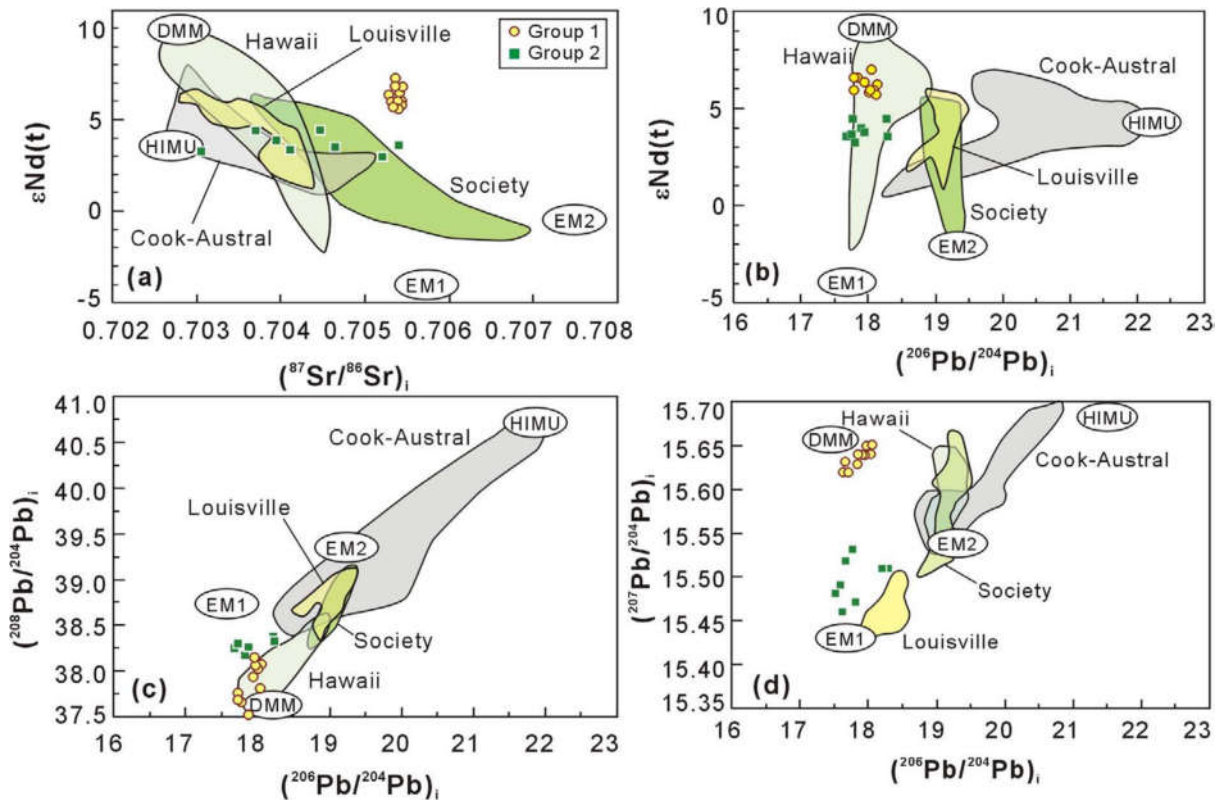


Fig. 8. Age-corrected Sr, Nd and Pb isotopic ratios of basalts in the Bayingou ophiolitic mélange. The coloured fields enclose present-day values for four other Pacific hot spot islands or island groups. Boxes marked DMM, HIMU, EM1, and EM2 indicate the postulated mantle end-member components of Zindler and Hart (1986). DMM—depleted mid-oceanic ridge basalt mantle, HIMU—high Th/U mantle, EM1—enriched mantle with low  $\epsilon\text{Nd}(t)$  and  $206\text{Pb}/204\text{Pb}$ , and intermediate initial  $87\text{Sr}/86\text{Sr}$ , EM2—enriched mantle with intermediate  $\epsilon\text{Nd}(t)$ , and high  $206\text{Pb}/204\text{Pb}$  and initial  $87\text{Sr}/86\text{Sr}$ . Data sources: See Fig. 7.

As mentioned above, the geochemical attributes of both groups imply no arc-related signatures, however, there are some contemporary arc-related magmas in North Tianshan accretionary complex (Wang et al., 2007; An et al., 2013; Xie et al., 2016). Furthermore, if both groups formed at oceanic spreading ridges, enriched basalts such as EMORB should also be found (Hemond et al., 2006; Nauret et al., 2006). However, there are no documented rocks with an E-MORB signature in this region. Therefore, we infer both groups were derived from a mantle plume, with varying degrees of previous melt extraction from their sources along with likely mixing between plume and MORB mantle sources.

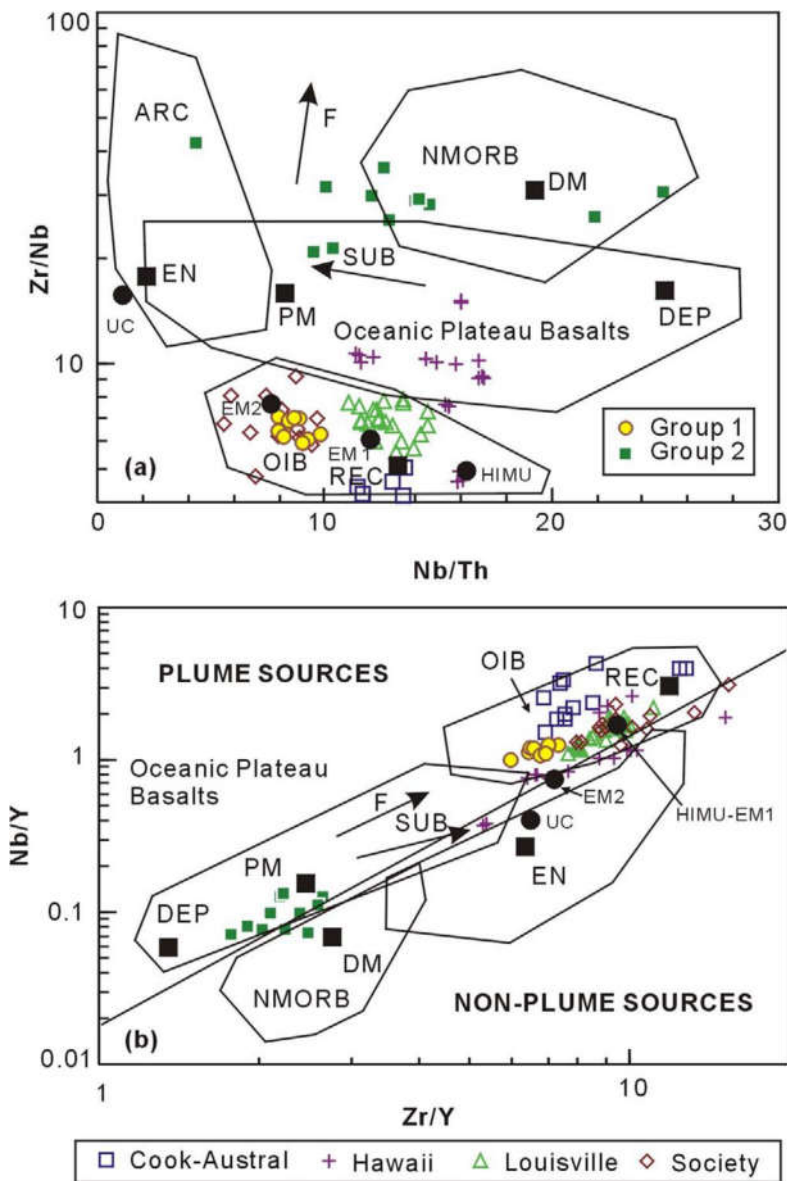


Fig. 9. Diagram showing mantle compositional components and fields for basalts from various tectonic settings (Condie, 2005). (a) Nb/Th versus Zr/Nb and (b) Zr/Y versus Nb/Y. Arrows indicate effects of batch melting (F) and subduction (SUB). Abbreviations: UC, upper continental crust; PM, primitive mantle; DM, shallow depleted mantle; HIMU, high Th/U mantle source; EM1 and EM2, enriched mantle sources; ARC, arc-related

basalts; NMORB, normal ocean ridge basalt; OIB, oceanic island basalt; DEP, deep depleted mantle; EN, enriched component; REC, recycled component. References: Weaver (1991), Condie (2003).

### 5.3. Seamounts in Junggar Ocean

The OIB-type pillow basalts that crop out in the Bayingou ophiolitic *mélange* of the North Tianshan occur in fault-bounded tectonic blocks within the *mélange* and are composed of massive basalts, gabbros, radiolarian cherts, pelagic limestones, basaltic breccias and tuffs. This ophiolitic *mélange* composition is similar to volcano-sedimentary sequences the accreted Paleogene ocean island sequences which are exposed in the Azuero Peninsula, west Panama (Buchs et al., 2011). Indeed, the association of basalt, breccia and limestone has been noted in many subduction–accretion complexes (e.g., Isozaki et al., 1990; Safonova and Santosh, 2014). Thus, we infer in conjunction with the geochemical signatures of the basalts, that the Bayingou ophiolitic *mélange* represents the remnants of a series of seamounts that are likely to have been oceanic islands or small plateaus. These fragments of seamounts are included in the imbricated thrust sheets and account for a relatively small volume of the *mélange* as the lowermost parts of the seamounts would have been subducted rather than

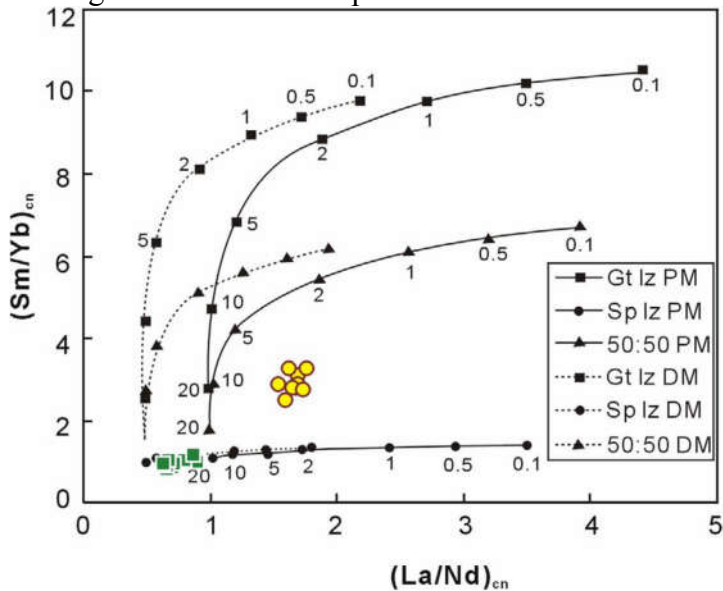


Fig. 10. Diagram showing the modelling results from pooled fractional melting calculations for various mantle source regions plotted on a  $(\text{La}/\text{Nd})_{\text{cn}}$  versus  $(\text{Sm}/\text{Yb})_{\text{cn}}$  diagram (after Kerr et al., 2015). Numbered ticks on the melting curves indicate percentage of partial melting. Gt, garnet; Sp, spinel; lz, lherzolite; PM, primitive mantle; DM, depleted mantle.

accreted during subduction–accretion processes (cf. Isozaki et al., 1990; Yang, 2016).

### 5.4. Geodynamic Implications for the Central Asian Orogenic Belt

The North Tianshan accretionary complex is one of the major accretionary complexes in the North Xinjiang region of western China and is located in the southeastern part of the Kazakhstan Orocline. The complex marks the final collision between the CTB in the south and the Junggar terrane in the north (Gao et al., 1997; Windley et al., 2007; Xiao et al., 2008; Zhu et al., 2009; An et al., 2013; Li et al., 2015). However, the timing of final subduction between CTB and Junggar terrane remains unclear. Xia et al. (2008) suggested that the complex formed in a rift setting in the Carboniferous due to the presence of so-called OIB-type volcanic rocks. However, Han et al. (2010) proposed that the collision occurred before the Late Carboniferous according to the presence of the Sikeshu ‘stitching pluton’ within the North Tianshan accretionary complex. Recently, Li et al. (2015) considered that the



subduction of the Junggar Ocean (a branch of the paleo-Asian Ocean) probably continued into the early Permian. Therefore, the accreted igneous rocks that it contains, such as the early Carboniferous Bayingou ophiolitic rocks of this study can shed new light on the evolution history of the North Tianshan accretionary complex.

This study, in conjunction with previous work (e.g., Wang et al., 2006; Han et al., 2010), supports the model that the Junggar Ocean, which was located between CTB and Junggar terrane, subducted southward beneath the CTB in the Late Devonian-Early Carboniferous. From this study it is clear that seamounts, comprising OIB-like pillow and massive basalts, dolerites and gabbros, developed on the ocean crust during its drift across a hotspot (Fig. 11a). As southward subduction proceeded, these upraised portions of oceanic crust were scraped-off and accreted in the wedge (Yang et al., 2015). Tectonic processes during accretion resulted in mixing of these fragments and the formation of an ophiolitic *mélange* in which the relicts of the seamounts are found as dismembered pieces (Fig. 11b). A recent review of OIB-bearing units in the CAO from the Late Neoproterozoic to Mesozoic by Safonova and Santosh (2014) suggested that the seamount-related magmatic activity was

### (a) Late Devonian-Early Carboniferous

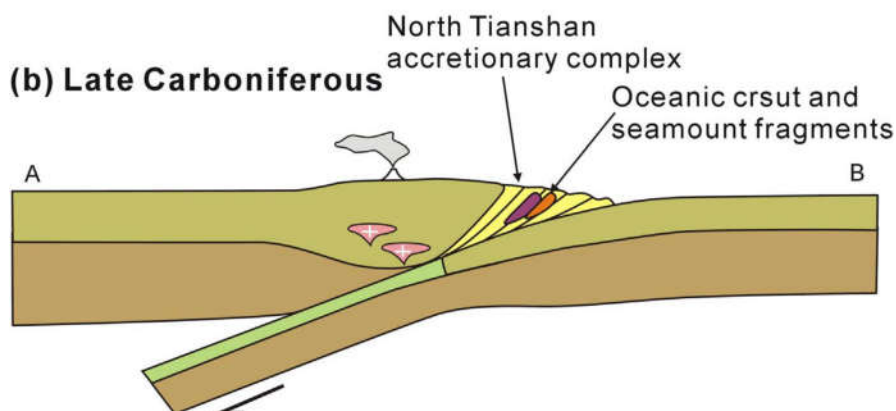
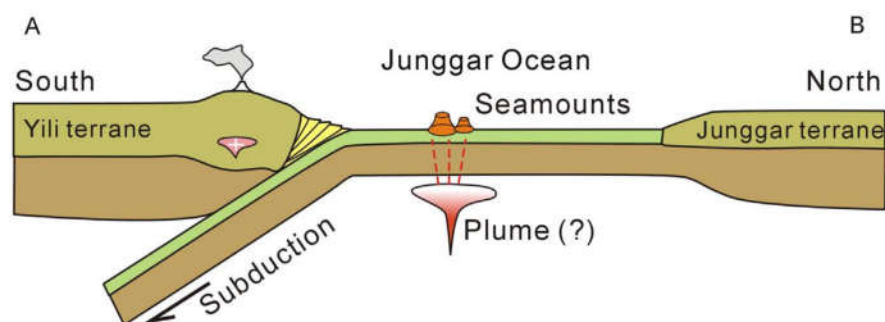


Fig. 11. Simplified geodynamic evolution of the Junggar Ocean in the Late Devonian-Late Carboniferous (after Han et al., 2010). Geodynamic reconstruction of the paleo-Asian Ocean based on Buslov et al. (2001). (a) Late Devonian- Early Carboniferous, the Junggar Ocean crust with seamounts were subducted southward beneath the Yili terrane. (b) In the Late Carboniferous, oceanic crust and seamounts were scraped-off and accreted in the accretionary complex resulting in the formation of the Bayingou ophiolitic *mélange*.

generally continuous during the development of the paleo-Asian Ocean (Yang et al., 2013, 2015). During the closure of the paleo-Asian Ocean, relict fragments of oceanic crust and seamounts were preserved as 'ophiolites' in the CAO accretionary complexes. In addition to the seamount fragments preserved in the ophiolitic *mélange*, the basement of the Tacheng block has recently been interpreted to represent a near-complete Early Paleozoic oceanic

plateau that was preserved in the accretionary complex of West Junggar (Zhang et al., 2017). It is widely accepted that there were multiple intra-oceanic subduction systems during the Paleozoic that closed the ocean basins which are now preserved in the CAO (e.g., Xiao et al., 2009). Therefore, the results of our study are consistent with the seamount accretion model to explain tectonic evolution of the CAO.

## 6. Conclusions

(1) The basaltic rocks in the Carboniferous Bayingou ophiolitic mélange show two compositional groups: one ocean island basaltlike (Group 1) and the other mid-ocean ridge-like (Group 2), that can be distinguished on the basis of their pillow structure and immobile trace element contents.

(2) The ocean island basalt-like Group 1 are derived from a deeper garnet-bearing lherzolite source with lower degrees (2–3%) of melting. In contrast the more trace element-depleted Group 2 basalts are derived from a shallower spinel lherzolite mantle source with ~15% partial melting. Paradoxically, Group 2, despite having more depleted trace element signatures than Group 1, has a more enriched isotopic signature, implying that the source of Group 2 was depleted by melt extraction shortly before formation. Both groups lie on a mixing line between DMM and EM1.

(3) Both groups of basalts in the Bayingou ophiolitic mélange represent the remnants of a series of seamounts, indicating that seamounts developed in the Junggar Ocean, as well as the paleo-Asian Ocean. Considering the spatial distribution of the Bayingou ophiolitic mélange and the North Tianshan accretionary complex, we suggest that the seamount accretion occurred during the tectonic evolution of the North Tianshan.

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